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J. F. B. Mitchell and T. S. Hills

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Meteorological Office (Met. O. 20)
London Road
Bracknell
Berkshire RG12 2SZ

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J F B Mitchell and T S Hills

Met O 20 (Dynamical
Climatology Branch)
Meteorological Office
London Road
Bracknell
Berkshire, UK RG12 2SZ

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SEA ICE AND THE ANTARCTIC WINTER CIRCULATION: A NUMERICAL EXPERIMENT

J F B Mitchell and T S Hills

Meteorological Office, Bracknell

Abstract

A numerical experiment has been conducted to test the sensitivity of a global general circulation model to changes in sea-ice extents in the Antarctic during winter. Three 112 day integrations have been made in which all the Antarctic sea ice poleward of 66°S was removed, commencing on the 10th June in the second, third, and fourth years of a control integration. There was a large increase in sensible heat flux over the anomaly, a warming over the Antarctic confined to the lowest atmospheric layer, and a 2 ms^{-1} reduction in the westerly flow around the periphery of the (new) sea-ice margin, in broad agreement with a previous study by I Simmonds. The increased heating over the anomaly was accompanied by a decrease in surface pressure which was not found by Simmonds, and possible explanations of this discrepancy are considered. The results are related to previous work on the effect of increased CO_2 using prescribed changes in sea surface temperature and sea-ice extents. The implications of the results for the parametrization of the heat flux through sea-ice cover, and for coupled ocean-atmosphere models are discussed.

Sea-ice and the Antarctic Winter Circulation: A Numerical Experiment

1. Introduction

Sea-ice has a profound effect on the transfer of energy between the atmosphere and the underlying ocean. The fluxes of sensible and latent heat into the atmosphere over sea-ice are substantially smaller than over the open ocean. The absorption of solar radiation by the surface is considerably reduced, as sea-ice has a much higher reflectivity than open water. The first process affects the atmosphere directly; the second influences it indirectly by changing the evolution of the surface. In this paper, we are concerned with the response of the atmosphere to given surface conditions, and so the second effect will not be considered further.

The largest fluxes of heat into the atmosphere in high latitudes in winter occur over open water. For example, Bunker (reported by Herman and Johnson, 1980) estimates a long term mean sensible heat flux of 35-70 Wm^{-2}, and a latent heat flux of 50-200 Wm^{-2} using ships' data from ice free areas of the Labrador Sea and Davis Strait in bulk aerodynamic formulae. Considerably higher fluxes are to be expected during outbreaks of cold air from the Arctic. In contrast, observations made in the Central Arctic (Vowinckel and Taylor, 1965; Vowinckel and Orvig, 1970) indicate that the turbulent heat flux between sea-ice and the atmosphere is small, typically 1-2 Wm^{-2}, though Fletcher (1965) derives an upward flux of sensible heat of 15 Wm^{-2}. According to Maykut (1976), leads of open water begin to dominate the large scale turbulent heat exchange where they account for more than 1% of the ice pack. However, it is generally accepted that less than 5% of the central Arctic is occupied by open water in winter, and observations during 1960 reported by Wittman and Schule (1966) indicated that less than

1.5% of the surface area was covered by leads. Thus, the flux of heat into the atmosphere increases by at least an order of magnitude as one goes from the central Arctic to the neighbouring ocean.

Given that anomalies in sea-ice cover tend to persist for several months (for example Lemke et al 1980), one might expect the associated anomalies in surface heat flux to influence the atmospheric circulation. In the Atlantic in winter, the climatological mid-latitude pressure trough lies parallel to and equatorward of the edge of the pack ice. There is a tendency for depression tracks and the position of the jetstream in the Atlantic sector to be further south in years of extensive cover than in years of relatively little sea-ice (for example Wiese 1924, Schell 1970, Ratcliffe and Morris 1978). Similar relationships have been reported in experiments with atmospheric general circulation models, using anomalies in ice cover from the last glacial maximum (for example, Williams et al, 1974), or corresponding to the difference between minimum and maximum extents observed in recent years (Herman and Johnson, 1978).

In the Antarctic, most of the pack ice melts in summer and reforms in winter. Little is known of the thickness distribution, though Weller (1980) assumed that the average thickness of inner pack ice is about 80 cm, considerably less than the 2 to 4 m indicated by submarine observations made in the central Arctic (Le Shack, reported in Hibler 1979). Attempts have been made to deduce the concentration of sea-ice cover from satellite data (for example Zwally et al 1979), and such evidence as is available suggests that in winter months, the ice is thinner and there is more open water than in the Arctic. Weller (1980) estimated a turbulent heat flux into the atmosphere of 117 Wm^{-2} in the inner zone (defined as where the concentration of sea ice cover exceeds 85%) and 189 Wm^{-2} in the outer zone

(ice concentration 15-85%), whereas over the open ocean near 50°S, the turbulent heat flux has been estimated at 60 to 116 Wm^{-2} (Zillman, 1972; Van Loon 1984), the highest values occurring in the Southern Atlantic Ocean. Although Weller claims no great accuracy for his estimates, it seems probable that the contrast in the turbulent heat flux from pack ice to open ocean is less pronounced than in the Arctic, or even reversed. The relatively large areas of open water, forced by the divergent pattern of surface stress, and the low relative humidity, produced by cold dry air flowing off the Antarctic continent (Andreas and Ackley, 1982) may be sufficient to produce the large fluxes over Antarctic sea-ice cited by Weller. If such fluxes are typical, the overlying atmosphere will become warmer than it would over the thicker, denser ice found in the Arctic. Consequently, when this warmed air flows off the Antarctic ice-pack, the flux of heat from the neighbouring ocean will be less than it would be into colder air moving off more compact ice cover. This contrast in the meridional variation in surface heating may be why, in winter, the Antarctic surface pressure trough, unlike its northern hemisphere counterpart, lies over the pack ice at most longitudes, particularly in the Ross and Weddell seas. Several authors (Ackley, 1981; Streten and Pike, 1980; Cavalieri and Parkinson, 1981) have considered the relationship between southern hemisphere sea ice extents and atmospheric circulation, though such investigations are inevitably hampered by the lack of reliable data.

Evidence concerning the Antarctic from numerical studies of the last glacial maximum is contradictory. Gates (1976) found that the southern circumpolar trough moved equatorward and was less intense in winter with increased sea ice cover, whereas Manabe and Hahn (1977) found that the

southern surface pressure trough shifted polewards. The statistical significance of these changes was not tested. However, in both experiments, substantial changes to sea surface temperatures, ice sheets and snow cover in the northern hemisphere may have modified the local response in the Antarctic. Simmonds (1981) reported a September simulation made using a hemispheric atmospheric general circulation model with reduced (March) Antarctic sea ice extents, and found surface pressure increased in high and middle latitudes. The simulated depression track did not shift south with the ice edge, as might have been expected for example, from the results of Herman and Johnson. Simmonds was unable to explain his results in physical terms, though he did note that the changes in the model's zonally averaged surface pressure were similar to those observed between September and March.

In view of the lack of observational data in southern high latitudes, and the inconclusive nature of numerical studies to date, we have performed a winter Antarctic sea ice anomaly experiment with the Meteorological Office 5-layer model, in which all Antarctic sea ice equatorward of 66°S was removed. The results have been analysed with the following objectives:

1. To find out if Simmonds's findings, based on a 130 day integration, would hold for a longer experiment with a more detailed atmospheric model.

2. To quantify the sensitivity of the model to the insulation of the atmosphere from the ocean by sea ice.

3. To assess the impact of prescribed changes in sea ice in a previous study of the effect of CO_2 and climate. (Mitchell and Lupton, 1984).

4. To estimate the likely effect on the atmosphere of errors in the simulated sea ice extents in coupled ocean-atmosphere models.

2. The model, model climatology and the experiments

a. The model

In this study we have used the Meteorological Office 5-layer general circulation model originally developed by Corby et al (1977). It is a primitive equation model, using an energy conserving finite difference scheme with a 10 minute time step. It uses a σ co-ordinate system (σ = pressure/surface pressure), with the atmosphere divided into 5 layers of equal mass. The horizontal grid gives a quasi-uniform resolution over the sphere with a grid length of approximately 330 km.

An explicit boundary layer height is carried and used to control convection and surface exchanges. Supersaturation is removed by condensing excess moisture as rain, and the static stability is controlled using a penetrative convection scheme. The surface fluxes of heat and moisture are computed using bulk aerodynamic formulae, with the surface exchange coefficients dependent on surface stability (stable or unstable) and type (sea or land and sea-ice), as outlined by Mitchell (1983). The original model has been modified to include an interactive radiation scheme, soil moisture and snow variables, and seasonal variation of sea surface temperatures, sea-ice extents, ozone concentrations, and zonally averaged cloud cover (Slingo, 1982). The radiation scheme is similar to that of Manabe and Strickler (1964). Solar radiation is absorbed and scattered by atmospheric gases (ozone, carbon dioxide and water vapour), clouds and the earth's surface. The reflectivity of the surface is 0.8 over sea-ice, (0.5 if the surface

temperature rises to freezing point) up to 0.6 over snow-covered land and 0.06 over the oceans. Longwave fluxes are calculated using the emissivity approximation. The heat flux through the ice is estimated by assuming conduction through uniform ice cover 2 m deep as in Manabe, 1969. The model loses mass at the rate of 0.01 mb d^{-1} due to rounding errors (see Mitchell, 1983). This does not affect the interpretation of the results. The surface pressure at each point has been enhanced in retrospect by the fixed percentage required to maintain a global mean surface pressure of 988 mbs. Surface pressure has been reduced to mean sea level in Figure 1 to allow comparison with climatological data. Elsewhere, differences in surface pressure rather than surface pressure reduced to mean sea level are shown, as this avoids spurious effects due to extrapolation under high topography such as Antarctica.

b. Model Climatology

The main features observed in the annual cycle of the general circulation are represented in the $3^1/_2$ year control simulation. The simulated Antartic circumpolar trough is 15 mb less deep and 6° of latitude further north than observed (Figure 1). Studies made with this model (Hills, 1982) and elsewhere (Manabe et al, 1979a) indicate that these errors are less pronounced when horizontal resolution is enhanced. However, in models with higher horizontal resolution, the northern mid latitude trough becomes excessively deep in winter, whereas the southern hemisphere trough is well simulated, suggesting that the dissipation of kinetic energy over land, particularly mountains, may be inadequately represented in the models.

c. The experiment

In the experiment, all Antarctic sea ice equatorward of 66°S was removed and replaced by water at 0°C (Figure 2). This change is broadly similar to that observed between September and March except between the Ross Sea and the Antarctic peninsula, where the seasonal change is substantially larger. (The mid-month mean latitude of the sea ice boundary in the control simulation is 61.4, 57.2 and 60.2°S in July, August and September respectively, whereas in March it is 67.3°S)

. Three 112-day anomaly integrations were made, commencing from data for 10th June in the second, third and fourth years of the control integration. The changes over the final 92 days (July, August and September) have been analysed. It has been assumed that 20 days is sufficient for the atmosphere to adjust to the perturbation, and this view is supported by the lack of any consistent trend in the response over the succeeding 3 months.

3. Results

a. Thermal Response

The immediate effect of replacing sea-ice with open water is to raise the surface temperature (Figure 3a). The increase in zonally averaged temperature reaches a maximum of over 20°C in the two most southerly rows from which ice was removed, and falls off sharply on either side. The maximum in the flux of sensible heat into the atmosphere near 52.5°S in the control integration shifts polewards to 64.5°S beside the new ice edge (Figure 3b). Thus there is a large decrease in the flux into the atmosphere centred near 52.5°S, and a massive increase in the two rows corresponding to the maximum

temperature increase (Figure 3c) which produces a warming of the overlying atmosphere (Figure 4). This atmospheric warming is largely confined to the lowest model layer as in the unperturbed simulation at these latitudes the vertical profile of temperature is extremely stable (Figure 4). Although there is a sharp maximum in the increase in sensible heat flux at 64.5°S, the atmospheric warming is distributed more evenly with latitude, due to lateral mixing by atmospheric disturbances (Figure 5a). This enhancement of atmospheric warming adjacent to the maximum contributes to the decrease in sensible heat flux on either side of the massive increase at 64.5°S (Figure 3c). The effect of the atmospheric warming on the zonal wind (Figure 5b) is discussed in the next sub-section.

The enhancement of surface temperature raises the saturated vapour pressure at the surface, and increases evaporation (Figure 3d). Again, there is a sharp maximum near 64.5°S, and a corresponding broad maximum in the change in specific humidity of the lowest atmospheric layer (not shown). As a result evaporation, like sensible heat flux, is reduced outside the region in which sea ice was removed. The globally averaged turbulent energy flux into the atmosphere increases by less than 1 Wm^{-2}.

b. Dynamical Response

The changes in the diabatic heating of the atmosphere (Figure 6a) are dominated by the transfer of sensible heat from the surface. This intensifies the mean upward motion near the ice edge, which is compensated at neighbouring latitudes by relative descent (Figure 6b), and is accompanied by a broader reduction in zonally averaged surface pressure (Figure 6c). The largest falls in surface pressure coincide

with the maximum increase in diabatic heating. This relationship is substantiated by the geographical distribution of the response in surface pressure, the largest falls occurring in the southern Atlantic and Indian Oceans (Figure 7a), where the changes in sea-ice cover (Figure 2) and hence sensible heating are largest, and rises occurring in the southeastern Pacific Ocean, where the distribution of sea-ice is largely undisturbed. The zonally averaged surface pressure trough both intensifies (by 3 mb) and moves about one grid length (3° of latitude) poleward (not shown).

The changes in surface pressure alone would produce and enhance easterly flow over the Antarctic continent, and strengthen westerly flow equatorward of the new ice edge. However, the warming of the lower atmosphere, and consequent raising of the heights of constant pressure levels over Antarctica weakens the westerly flow around the continent in all but the lowest model layer (Figure 5b).

4. Discussion

a. Comparison with Simmonds's (1981) Experiment

The perturbation considered by Simmonds (1981) was slightly larger than that considered here, especially in the sector from the Ross Sea to the Antarctic Peninsula, where the ice edge was moved from about 62° to 71°S. The resulting changes in turbulent heat flux were similar to the present experiment, but the changes in atmospheric temperature were smaller and largely confined to the region of the ice anomaly. This produced a weakening of the upper westerly flow equatorward of the anomaly as in the present experiment, and strengthened the westerly flow over the Antarctic as one would expect

from the thermal wind approximation. (The increased westerly flow over the anomaly in the lowest model layer is associated with the changes in meridional surface pressure gradient).

In both experiments, surface pressure increases over eastern Antarctica and southwest of Tierra del Fuego, and decreases in the southern Atlantic, and around the periphery of the Antarctic from South America eastwards to the Indian Ocean. However, from the southern Indian Ocean to the Ross Sea, Simmonds found surface pressure rose (Figure 7b), whereas in the present study, there is a narrow region of reduced pressure over the ice anomaly. It is possible that a spectral model retaining only 15 waves is incapable of resolving such a small feature. The discrepancies are still apparent when just the changes in September (the period chosen by Simmonds, not shown) are considered, except there is closer agreement southwest of Australia and eastern New Zealand. In Simmonds' model, the drag coefficient used in the calculation of surface exchanges has the same value over sea-ice and ocean points. In the present model, it depends on both surface type and low level static stability (see Corby et al, 1977), so that in changing from stable conditions over sea-ice to unstable conditions over the ocean (Figure 4), the drag coefficient is increased from 0.3×10^{-3} to 2.0×10^{-3}, and the turning of the wind within the boundary layer is reduced from $\pi/6$ to zero. An increase in the drag coefficient will increase the surface drag in the lowest layer, tending to fill centres of low pressure. However, the decrease in the turning of the wind will have the opposite effect, tending to reduce low level mass convergence into depressions, so the net effect on the depth of the Antarctic surface pressure trough is

uncertain. Hence, the difference in the surface parametrisations used in the two models may or may not contribute to the differences in the models' responses.

We have performed a Student's t-test on the changes, assuming as did Simmonds that successive monthly means are independent. The reductions in surface pressure over the anomaly are mostly significant at the 95% level of confidence except near the Ross Sea and west of the Antarctic Peninsula, where the change in sea-ice distribution was minimal (Figure 7c). There are also substantial areas in the tropics where there is a small but significant increase in surface pressure. (The longitudinal bands are due to the presence of weak two grid length waves in the surface pressure field which attain slightly different amplitudes in the control and anomaly integrations). The response in our experiment is remarkably consistent, with reduction in zonal mean surface pressure occurring over the region where sea-ice was removed in all but one of the nine months of the experiment (Figure 8), and even then, the sea level pressure changes are locally a minimum in that region. In view of this consistency, and because we have been able to explain the simulated response on physical grounds, it is possible that the sea level pressure changes in Simmonds's experiment arose largely through chance. Although there were large coherent areas in Simmonds' results, this may have been due to spatial correlations inherent in meteorological fields (Livezey and Chen, 1983), rather than a systematic response to the change in the surface boundary conditions. The area of significant change in the region of the anomaly was small compared with the scale of the changes in sea-ice cover.

b. Model Sensitivity to the Heat Flux through Sea Ice

The heat flux through sea ice in this (and other) atmospheric models is estimated assuming conduction through ice 2 metres thick with complete cover. This may be acceptable for much of the Arctic, where, as noted earlier, the ice thickness is typically 2 to 4 metres. (The use of a lower than average observed value in the model allows implicitly for some heat transfer to the atmosphere from leads). However, the assumed ice depth is certainly unrepresentative of the Antarctic, where the ice is generally less than one metre thick, and there is more open water due to the divergent pattern of surface stress. Since, on the other hand, we have completely removed the insulation due to sea ice in our anomaly experiment, we have an estimate of the upper limit of the error attributable to underestimating the heat flux through sea ice.

In the control simulation the turbulent heat flux around the periphery of the Antarctic is larger than the climatological estimates, averaging 130 Wm^{-2} near 50°S in July, for example, whereas near 65°S over sea-ice, the turbulent flux of heat is small and directed downwards. In contrast, Weller (1980) estimates large upward fluxes there. In the anomaly integration the turbulent fluxes at 50°S and 65°S in July are 94 Wm^{-2} and 201 Wm^{-2} respectively, in closer agreement with climatological estimates, although the latter figure is probably excessive. The error in the depth and position of the zonally averaged trough has been reduced by 3 mbs and 3° of latitude respectively. It should be recognised that errors in the depth and position of the circumpolar trough may also be reduced by enhancing the horizontal resolution of the model (as noted in section

2). Nevertheless, it is likely that the depth of the surface trough is underestimated in climatological data as the observational network is sparse (Zillman, 1983).

c. Effect of prescribed changes on sea ice in a previous study on CO_2 and climate

Mitchell and Lupton (1984) carried out an experiment in which CO_2 concentrations were quadrupled and changes in sea surface temperature and sea ice were prescribed. The increments to sea surface temperature were derived from previous integrations, but the reduction in sea-ice extents was made on a more arbitrary basis. In the southern hemisphere during winter the sea ice boundary was moved 6° of latitude poleward, which is broadly similar to the change in the present experiments. The zonally averaged June to August changes in turbulent heat flux and surface pressure are qualitatively similar, but are smaller than in the present experiment particularly in the region of the anomaly. The increase in atmospheric temperature over Antarctica and the consequent weakening of the westerly flow around the continent are much more pronounced in the earlier experiment, presumably due to the increases in CO_2 and sea surface temperature.

Comparison with studies of increased CO_2 made using low resolution atmospheric models coupled to a simple oceanic mixed layer is hampered by the wide variety of sea-ice extents produced in the control simulation by such models, and the reluctance of investigators to report changes in atmospheric circulation. Thus, for example, Manabe and Stouffer (1980) grossly underestimate the extent of Antarctic sea-ice in the control simulation, whereas Hansen et al (1984) used a model which produces more realistic sea-ice extents:

neither of these authors report changes in circulation with increased CO_2. Washington and Meehl (1984), using a model that produces excessive sea-ice cover, find that on doubling CO_2 Antarctic ice extents recede by about 5° of latitude, similar to the change in the present experiment. During June to August, there are increases of up to 13 K in the temperature of the lowest atmospheric layer and a decrease of several metres per second in the westerly component of wind near 200 mbs in the region where sea-ice is removed. Mitchell et al (1986) also find that on doubling of CO_2, Antarctic ice extents recede by 5° of latitude, and that westerly flow decreases by 1 to 2 ms^{-1} and low levels warm by up to 12 K in the region where sea ice is removed. The zonally averaged changes in sensible heat flux are small (< 10 Wm^{-2}) at the latitudes where sea-ice is removed, and there is no evidence of a local reduction in surface pressure. Nevertheless, the strong surface warming and reduced tropospheric flow found in these CO_2 studies are consistent with the present work.

d. Coupled Ocean-Atmosphere Models

It has proved difficult to produce an accurate simulation of sea-ice extents in coupled ocean-atmosphere models (Manabe et al 1979b, Washington et al 1980), especially where oceanic dynamics have been ignored (for example, Manabe and Stouffer, 1980; Washington and Meehl, 1984). The present study indicates that errors in sea-ice extents may distort the atmospheric circulation, at least in winter. The simulation of sea-ice is extremely sensitive to errors in surface albedo, particularly in spring when sea-ice is most extensive, and in summer, when insolation is strongest. Anomalies generated in summer may persist well into winter, given the large thermal inertia of the

oceanic mixed layer. It is unlikely that the enormous heat fluxes into the atmosphere depicted in Figure 3 could be sustained for long periods, unless the ocean circulation adjusted so as to preserve the anomaly. For example, strong cooling in the region of the anomaly might induce sinking motion, and draw in warmer surface water to maintain the anomaly.

As noted in section 2b, atmospheric models with insufficient horizontal resolution underestimate the depth of the Antarctic circumpolar trough, and hence the westerly surface stress in southern mid-latitudes. When coupled to an ocean model, the resulting equatorward Ekman transport is underestimated, the simulated southern ocean temperatures become warmer than observed, and Antarctic Sea extents are reduced (Washington et al 1980). Bryan and Lewis (1979) suggest that the resulting reduction in meridional temperature gradient would further weaken the westerly flow, providing a positive climatic feedback. The results from the present experiment indicate that although the westerly flow in the middle and upper troposphere is weakened where sea-ice is removed, the changes near the surface are small or even reversed (Fig. 5b), indicating that the feedback proposed above may be weak.

5. Conclusions

The removal of sea-ice equatorward of 66°S in a winter simulation produced a large flux of sensible heat into the lower atmosphere over the anomaly. The atmospheric warming spread polewards over the Antarctic and weakened the westerly flow around the continent in all but the lowest model layer. These changes are in broad agreement with a similar experiment performed by Simmonds (1981), using a lower resultion hemispheric model.

However, in the present work there was a reduction of surface pressure in the region where sea-ice was removed and atmospheric heating was enhanced, where Simmonds found marked increases over much of this region. This discrepancy may have occurred because Simmonds's model had insufficient horizontal resolution to respond to the small scale forcing from the anomaly, or because of differences in the prescription of surface drag coefficients, or have arisen by chance given the short duration of Simmonds's experiments. The atmospheric response is consistent with that in the winter Antarctic in an earlier simulation with prescribed changes in CO_2, sea surface temperatures and (reduced) sea-ice extents (Mitchell and Lupton, 1984).

Zonally averaged surface pressure decreases by 7 mb in the region of the anomaly and the high latitude trough shifts about 3° of latitude poleward. As the present model formulation probably underestimates the heat flux through sea-ice and leads in the Antarctic, these changes may be regarded as an upper bound to the errors due to the parametrisations of this process. Finally, the experiment gives some indication of the likely effect of errors in simulated sea-ice extents on the overlying atmospheric circulation in coupled ocean atmosphere models. Although strong surface cooling will tend to re-form sea-ice where it has been removed, the persistence of sea-ice anomalies, particularly if reinforced by changes in ocean circulation, may be sufficient to produce large errors in such models.

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Zwally, H J; Parkinson, C; Carsey, F; Gloerson, P; Campbell, W J and Ramseier, R O 1979 Antarctic sea-ice variations 1973-5, NASA Weather Climate Review Paper No 56, 335-340.

Figure Captions

1. Pressure at mean sea level in the control integration. (Three year average, July to September). Contours every 4 mbs, stippled where less than 1000 mbs. As the model loses mass at the rate of 1 mb every 100 days (due to rounding errors) the model values have been adjusted at each point by a fixed percentage to preserve the initial global mean surface pressure.

2. Model sea-ice extents and changes in sea-ice (light stippling) for mid-August. Observed minimum ice extents are indicated by heavy stippling.

3. (a) Changes in surface temperature (°C) due to removing Antarctic sea-ice equatorwards of 66°S, (three year average over July, August and September, (b) Sensible heat flux (Wm^{-2}) in the control (solid line) and anomaly (dashed line) experiments, and changes in (c) sensible heat and (d) latent heat, (Wm^{-2}) for the same period. The wedge near 60°S indicates the latitudes over which sea-ice was removed.

4. Vertical profile of zonally averaged atmospheric and surface temperatures. (Three year average over July, August and September) solid line, control; dashed line, anomaly experiment.

5. Changes in zonally averaged atmospheric parameters (anomaly-control). Areas of decrease are stippled.

   (a) Temperature. Contours every 2.5°C, with a dashed contour at $\pm$1°C.

   (b) Zonally averaged west to east wind component. Contours every 1 m/s.

6. Changes in diabatic heating (Wm^{-2}), vertical velocity averaged through the atmosphere (nb s^{-1}) and surface pressure (mb), due to removing Antarctic sea-ice equatorwards of 66°S. The wedge near 60°S indicates the latitudes up to which sea-ice was removed.

7. (a) Changes in surface pressure due to removing Antarctic sea-ice equatorward of 66°S. Three year average over July to September. Areas of decrease stippled, with contours every 2 mb, and dashed contours at ±1 mb.

(b) Changes in mean sea level pressure due to reducing September sea-ice extents (Simmonds, 1981). The contour interval is 2 mb. Negative areas are stippled. Regions of differences significant at the 90% confidence level are enclosed by dashed contours.

(c) Student's t-statistic of changes surface pressure, using nine monthly data sets, assumed independent. Lightly stippled where decreases significant at the 90% level of confidence, heavily stippled where increases are significant (2-tailed test). Contours indicate 90%, 95% and 99% levels of confidence. For convenience, the contour of zero change has been included in the Figure.

8. Changes in zonally averaged surface pressure (mb) due to reducing Antarctic sea ice extents. Solid line, year 1; dashed line, year 2; dotted line, year 3. (a) July, (b) August, (c) September.

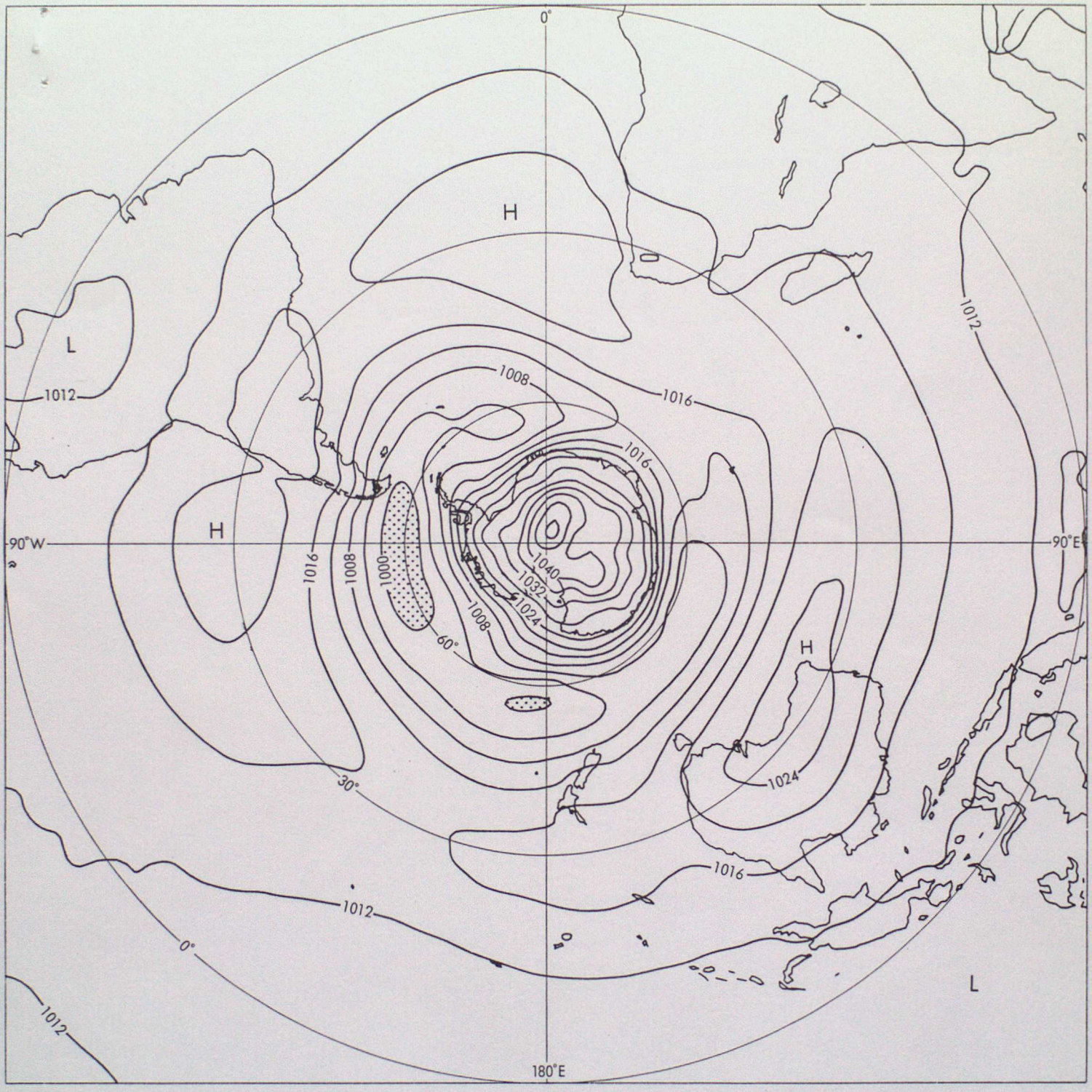

0°
H
L
1012
1008
1016
1016
1012
90°W
90°E
H
1016
1008
1000
1040
1032
1024
1008
60°
H
1024
30°
1016
1012
0°
1012
L
180°E

Figure 1

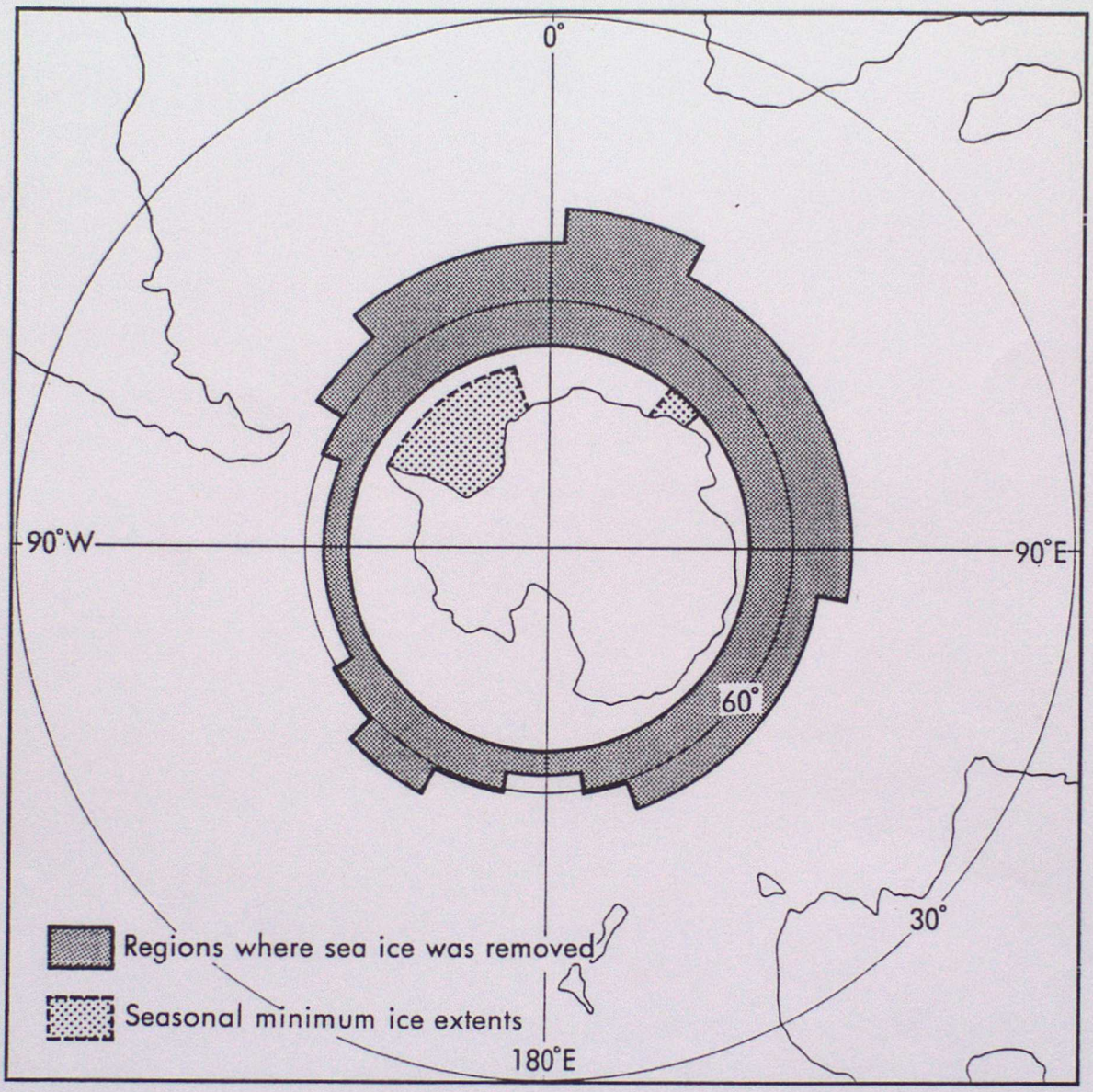
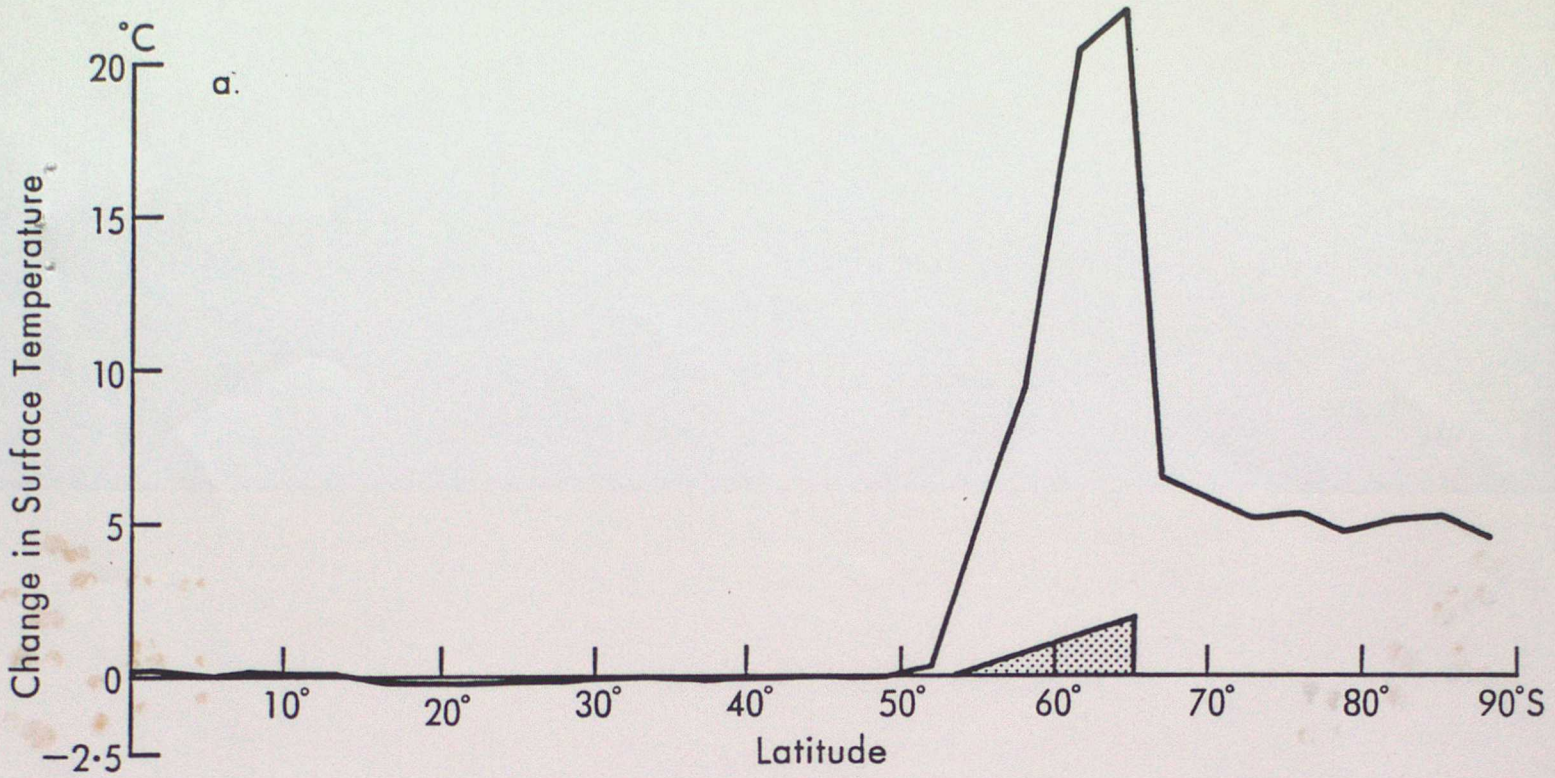

0°
90°W
90°E
60°
30°
180°E
Regions where sea ice was removed
Seasonal minimum ice extents

Figure 2



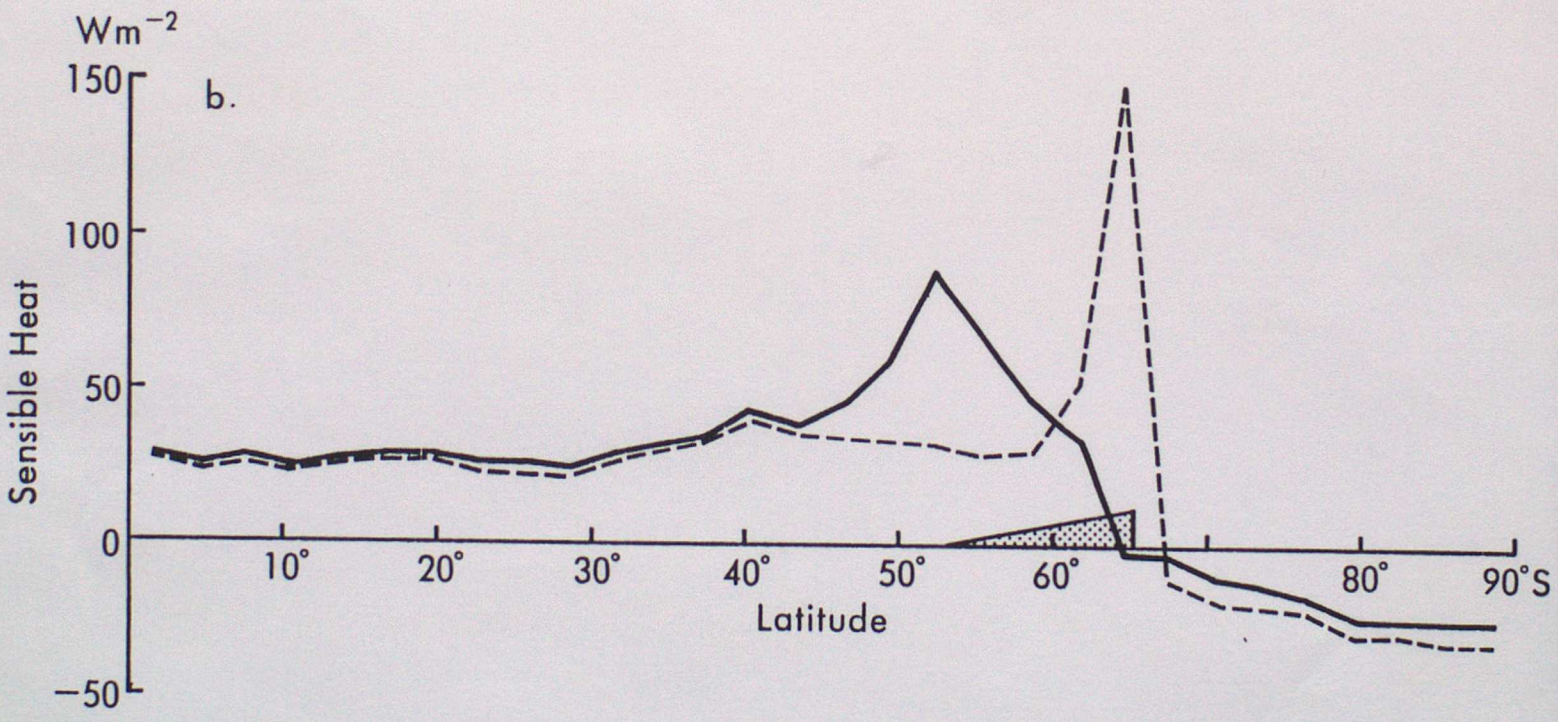


Figure 3a and 3b

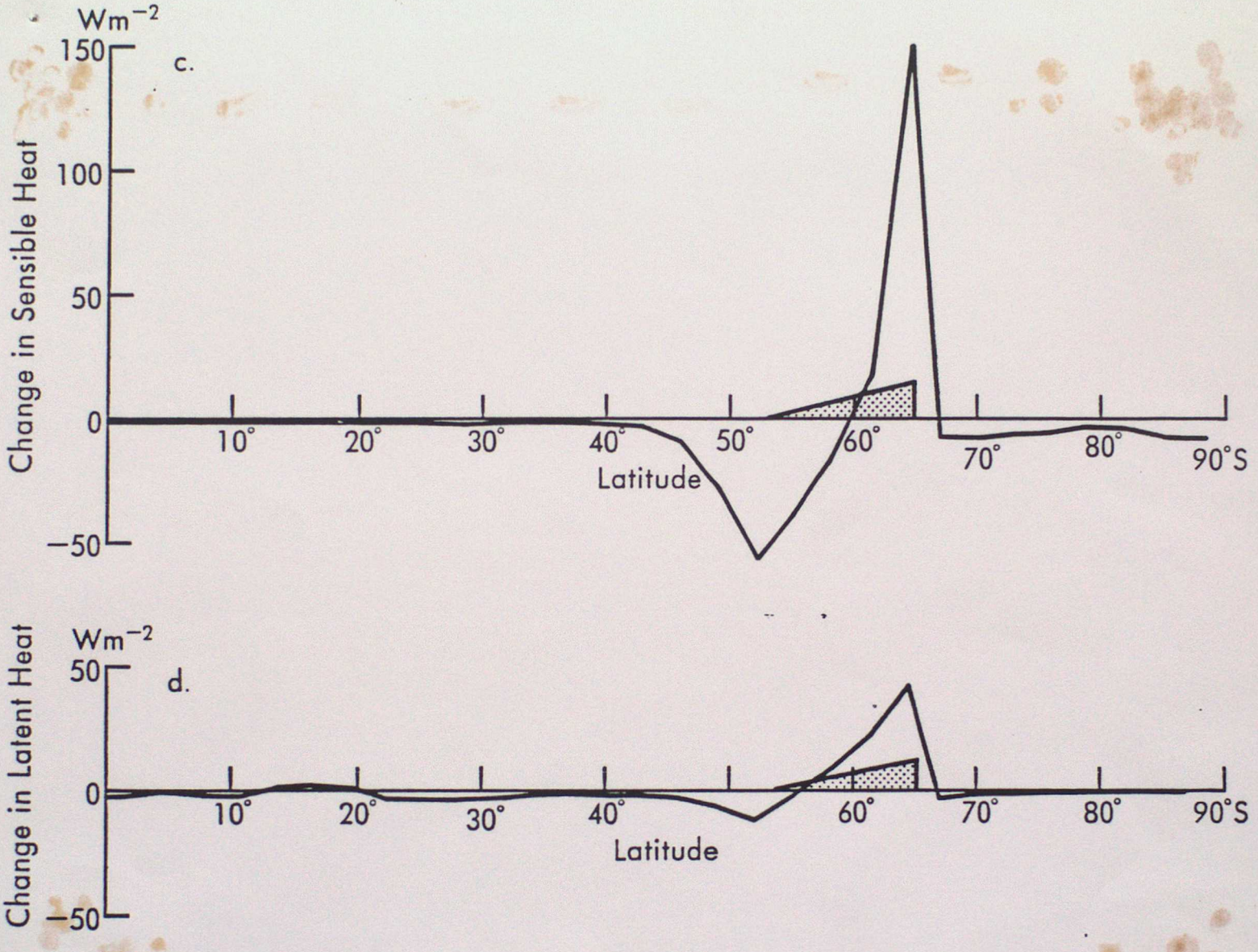
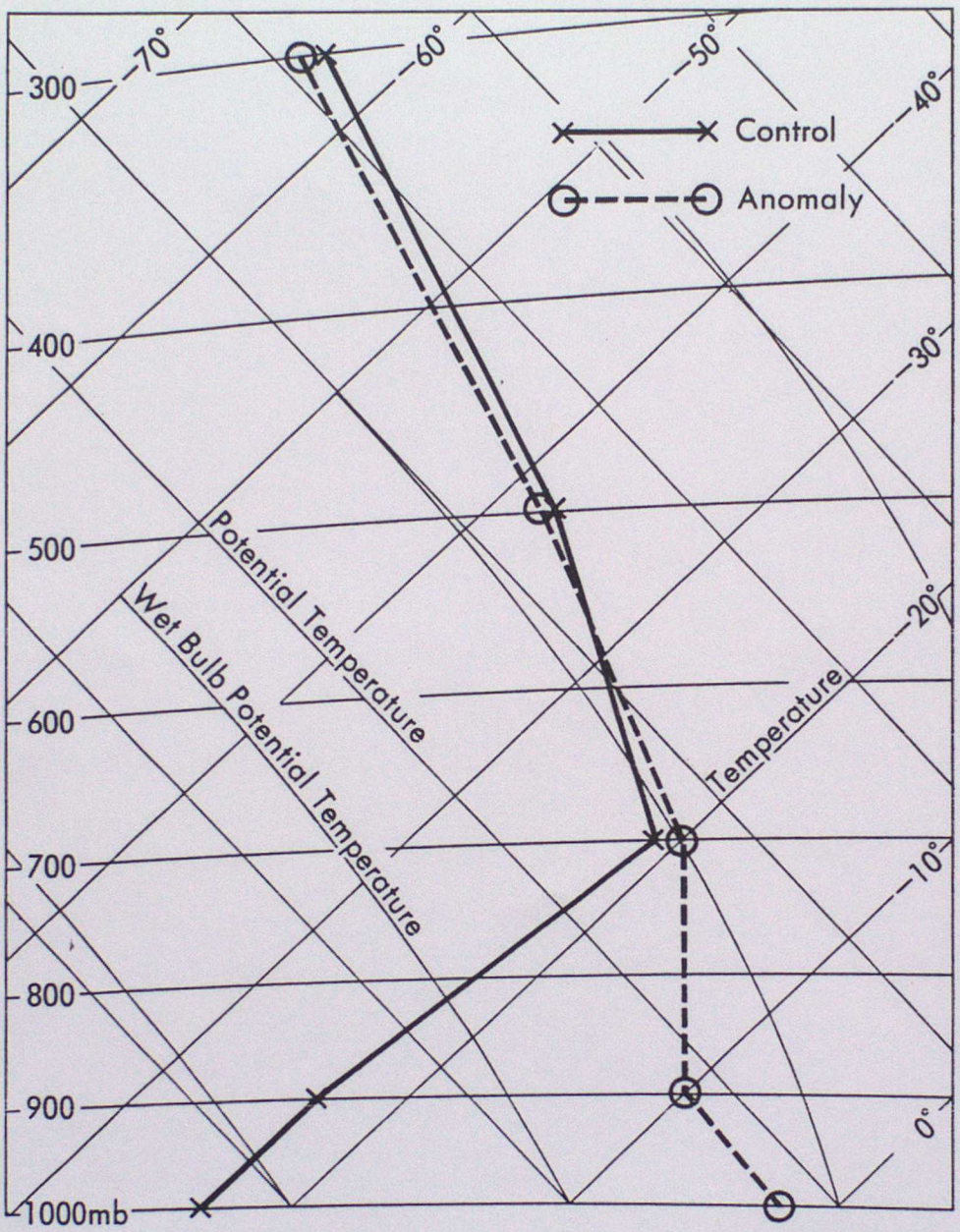


Figure 3c and 3d



Tephigram of zonally averaged Vertical Profiles at 64·5°S (July to September)

Figure 4

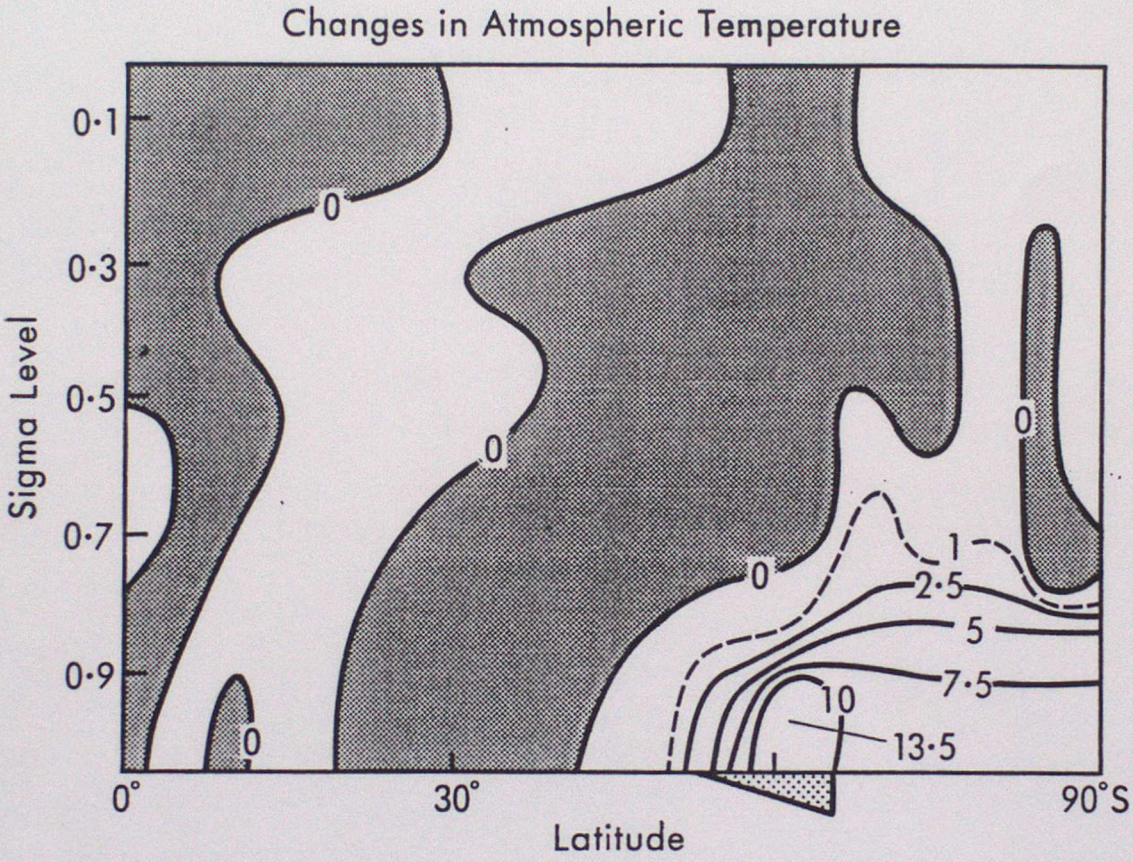

Changes in Atmospheric Temperature
Sigma Level
0·1
0·3
0·5
0·7
0·9
0
0
0
0
1
2·5
5
7·5
10
13·5
0°
30°
90°S
Latitude

Figure 5a

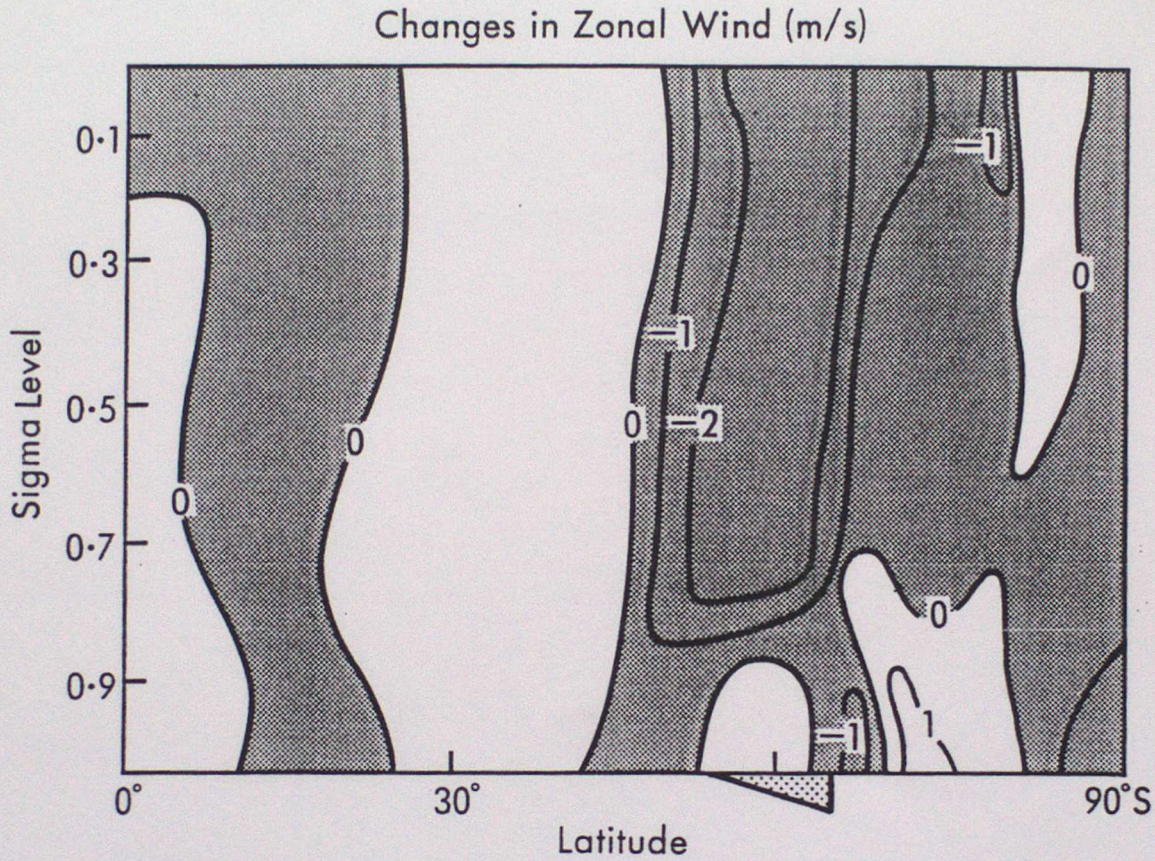

Changes in Zonal Wind (m/s)
Sigma Level
0·1
0·3
0·5
0·7
0·9
0°
30°
90°S
Latitude
0
0
0
0
0
0
−1
−1
−2
−1
1

Figure 5b

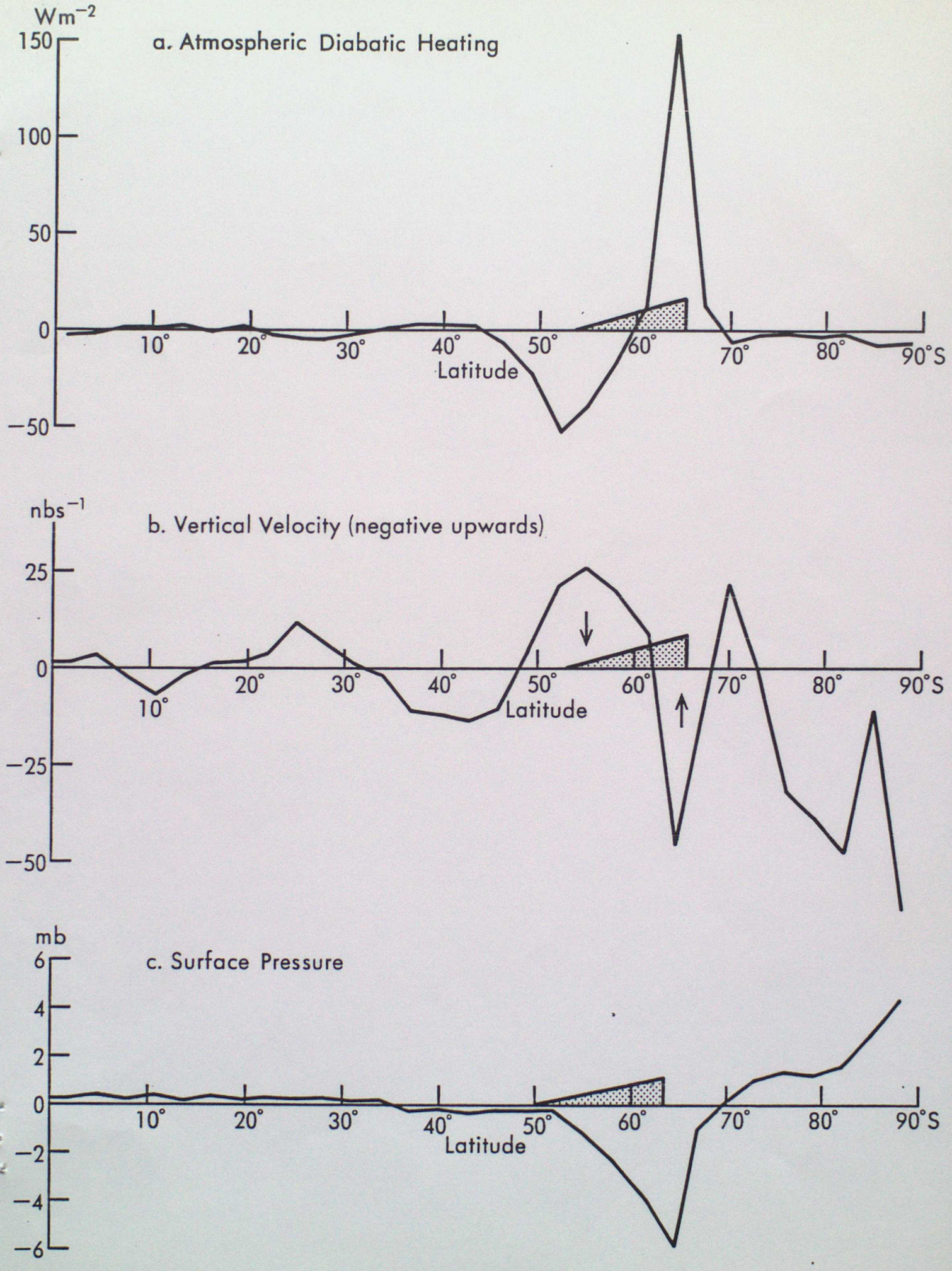

Wm^{-2}
a. Atmospheric Diabatic Heating
150
100
50
0
−50
10°
20°
30°
40°
50°
60°
70°
80°
90°S
Latitude
nbs^{-1}
b. Vertical Velocity (negative upwards)
25
0
−25
−50
10°
20°
30°
40°
50°
60°
70°
80°
90°S
Latitude
mb
c. Surface Pressure
6
4
2
0
−2
−4
−6
10°
20°
30°
40°
50°
60°
70°
80°
90°S
Latitude

Figure 6

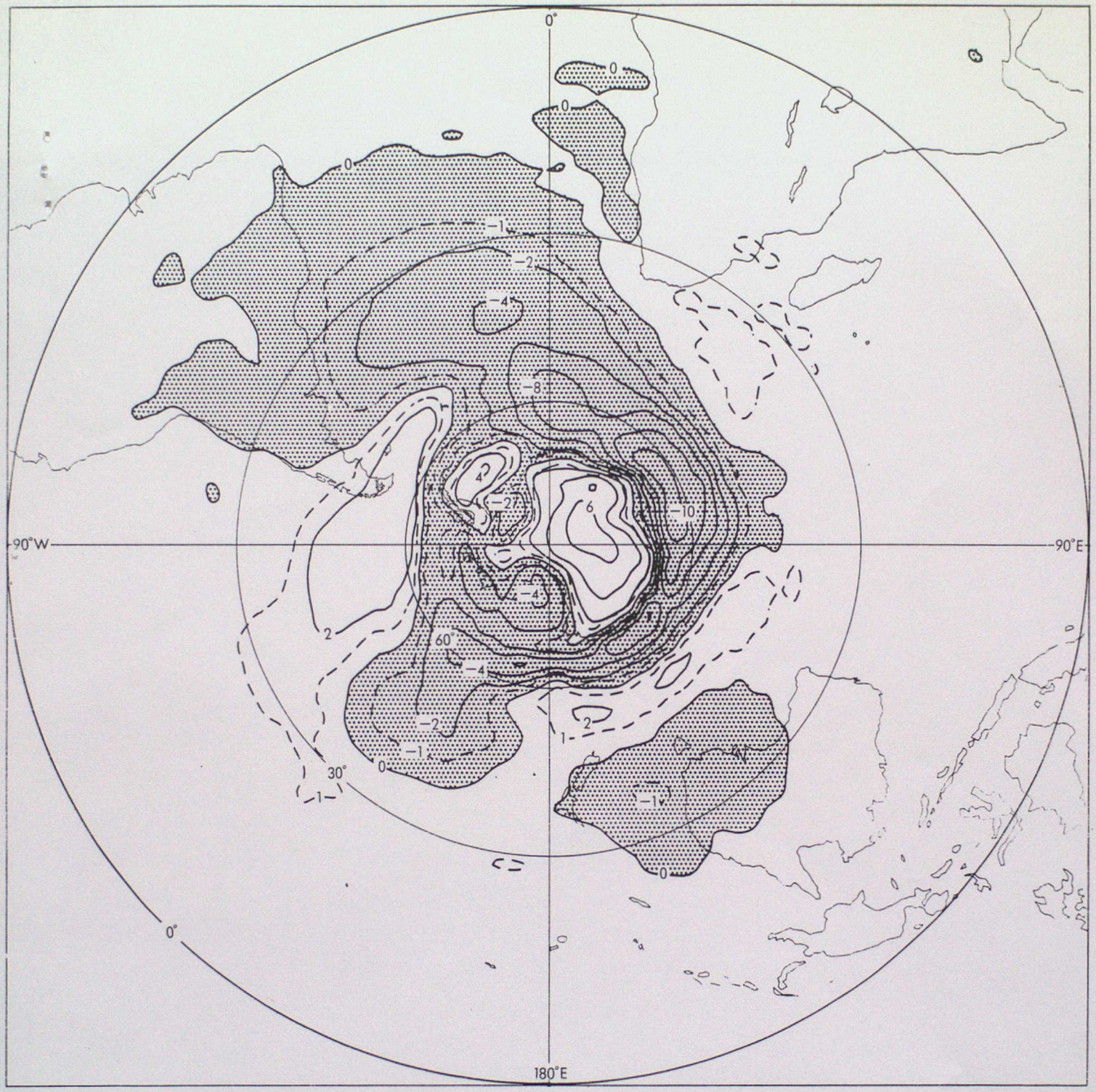


Figure 7a

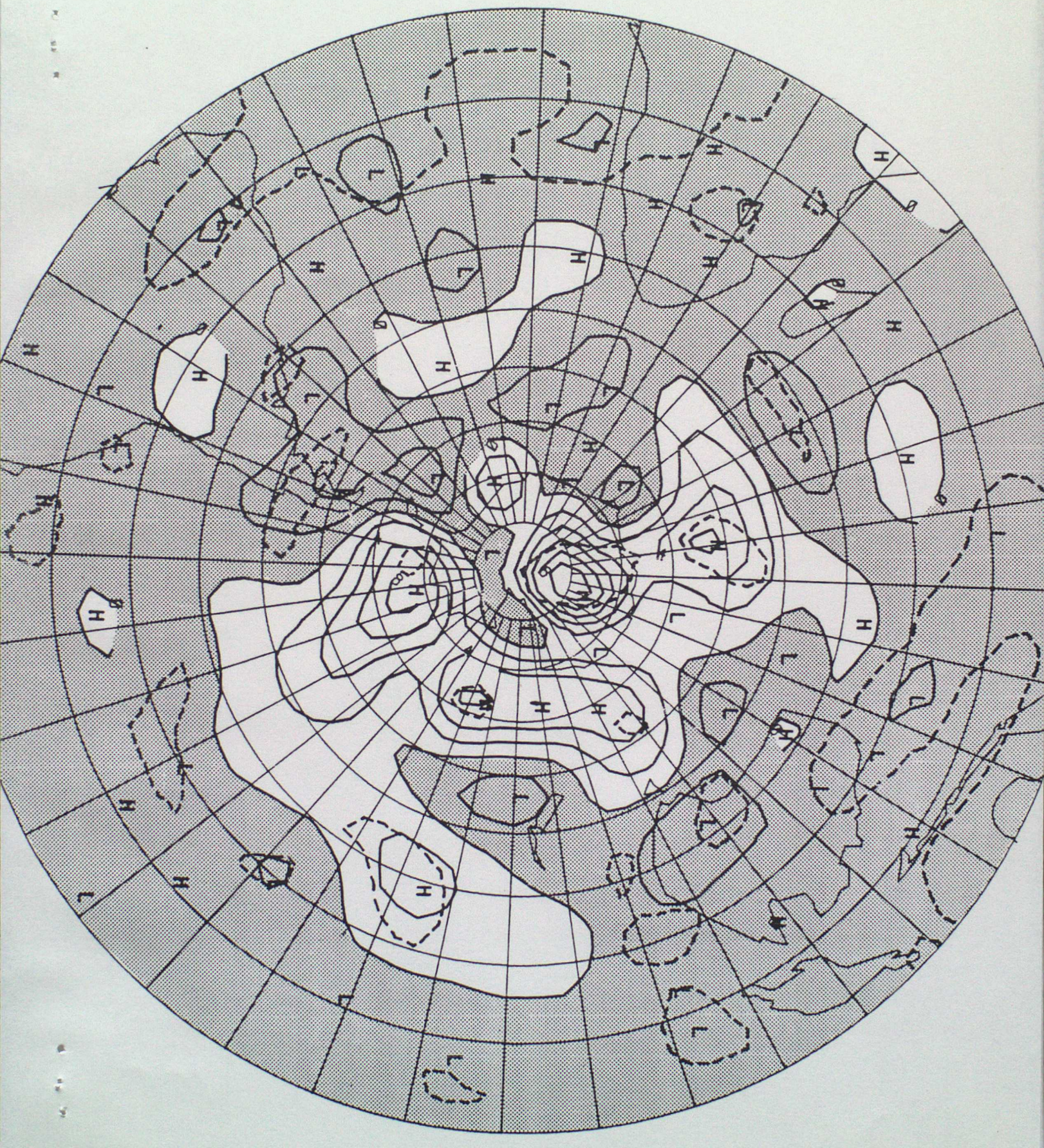


Figure 7b

Figure 7c

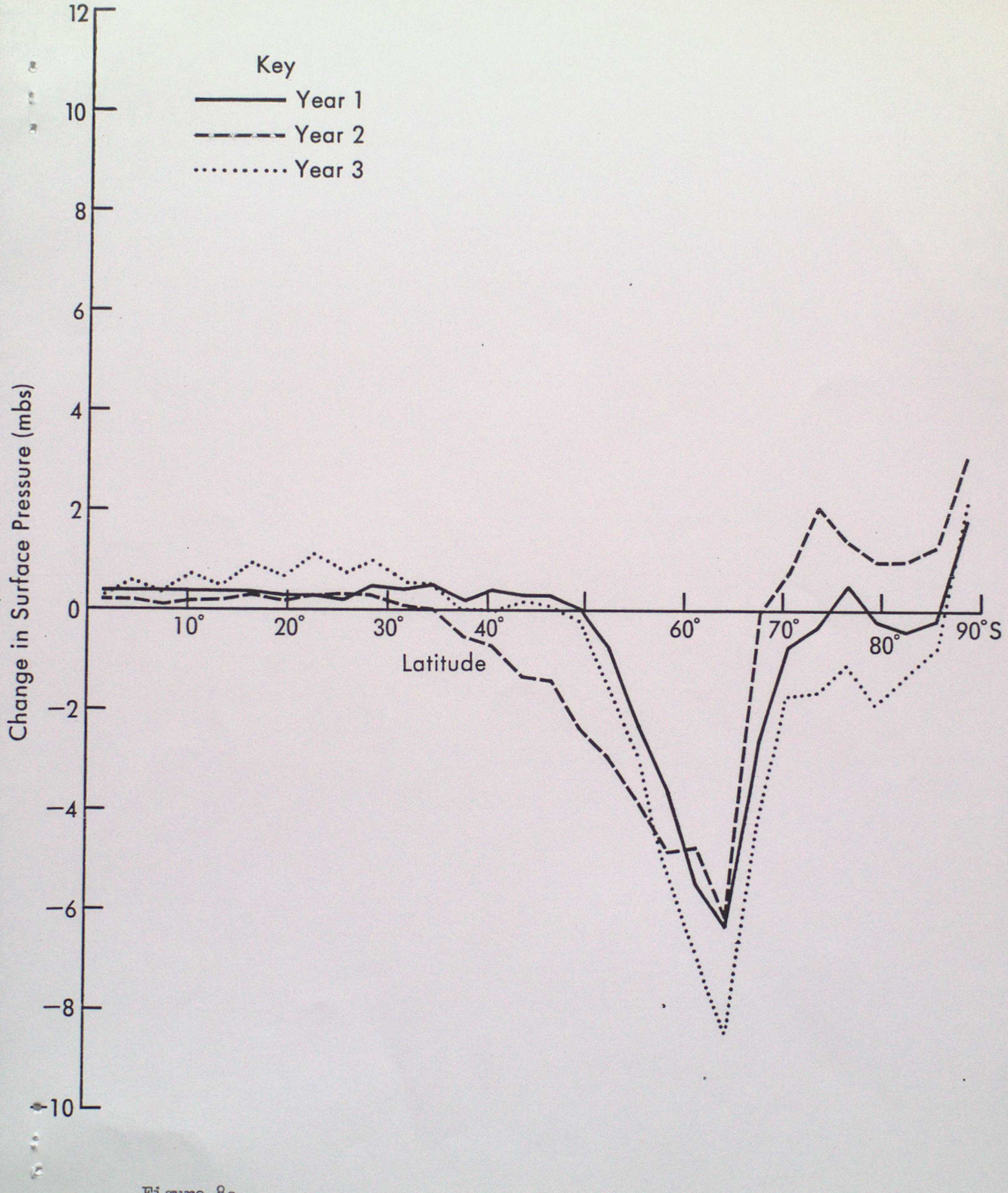

Key
Year 1
Year 2
Year 3
12
10
8
6
4
2
0
−2
−4
−6
−8
−10
Change in Surface Pressure (mbs)
10°
20°
30°
40°
60°
70°
80°
90°S
Latitude

Figure 8a

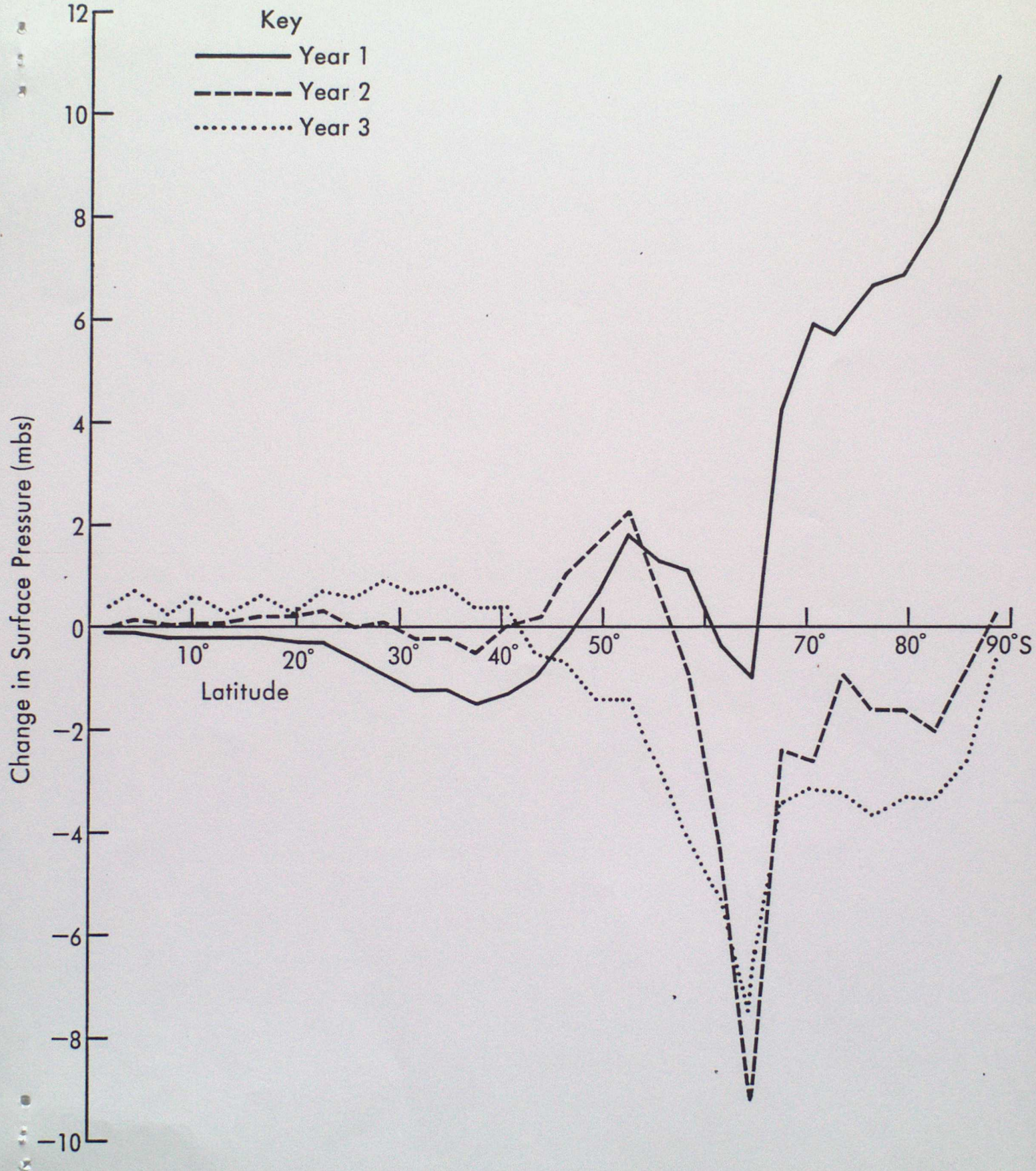

Key
Year 1
Year 2
Year 3
12
10
8
6
4
2
0
−2
−4
−6
−8
−10
Change in Surface Pressure (mbs)
10°
20°
30°
40°
50°
70°
80°
90°S
Latitude

Figure 8b

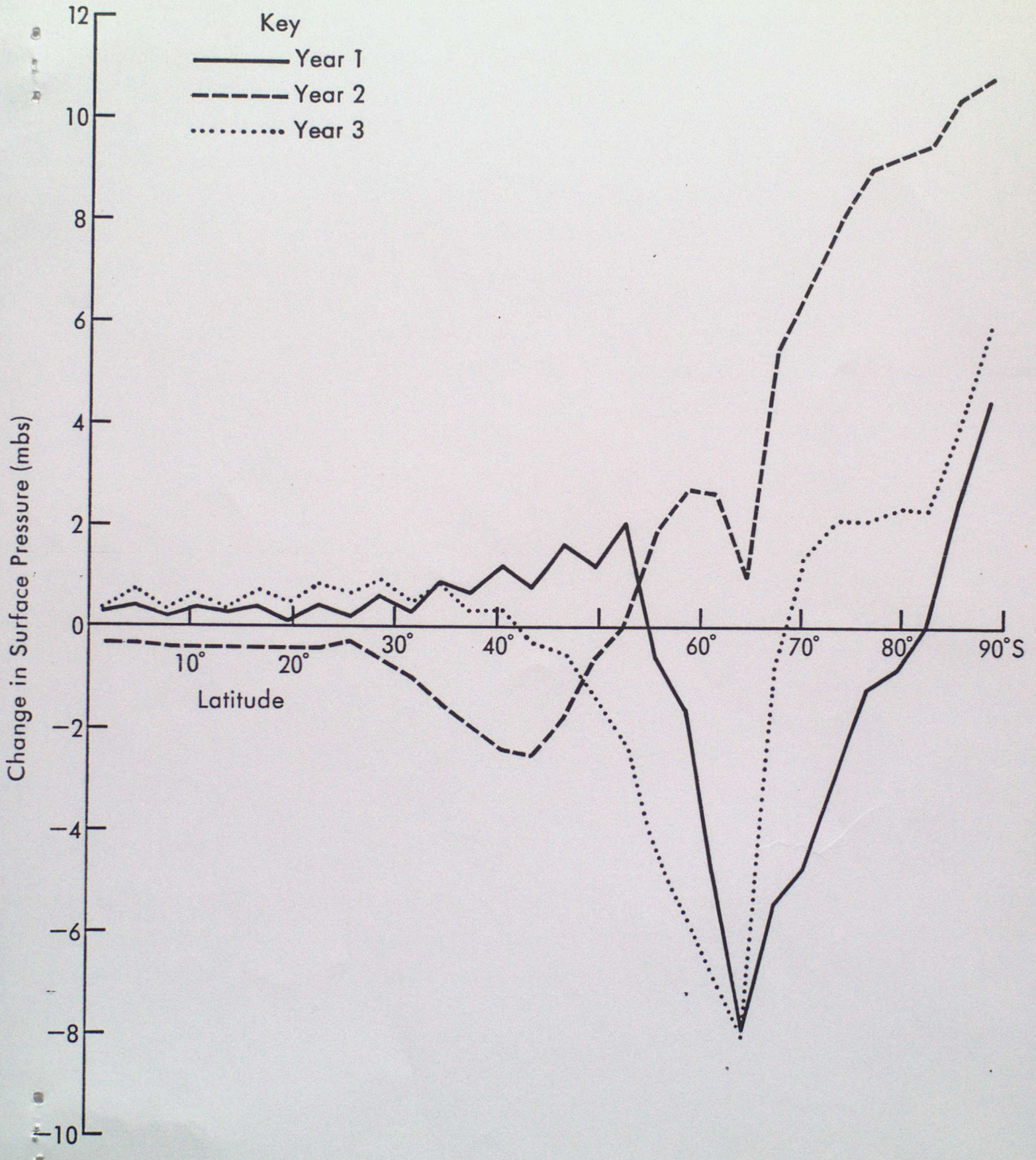

Key
Year 1
Year 2
Year 3
12
10
8
6
4
2
0
−2
−4
−6
−8
−10
Change in Surface Pressure (mbs)
10°
20°
30°
40°
60°
70°
80°
90°S
Latitude

Figure 8c

DYNAMICAL CLIMATOLOGY TECHNICAL NOTES

DCTN 1	June 1984	A scheme for incorporating the thermodynamic sea-ice model into a coupled ocean-atmosphere model. C Gordon and M Bottomley
DCTN 2	Sept 1984	The parametrization of the upper ocean mixed layer in coupled ocean/atmosphere models. C Gordon and M Bottomley
DCTN 3	Sept 1984	The response of a simple thermodynamic sea-ice model to GCM forcing appropriate for perennial sea-ice. M Bottomley
DCTN 4	Sept 1984	Potential vorticity in the stratosphere derived using data from satellites. S A Clough, N S Grahame and A O'Neill
DCTN 5	Sept 1984	Simulation of the seasonal cycle of SST using globally specified climatological data to force simple ocean models. N S Grahame
DCTN 6	Sept 1984	Development of an improved GCM radiation scheme. A Slingo and R C Wilderspin
DCTN 7	Sept 1984	On the specification of surface fluxes in coupled atmosphere/ocean general circulation models. J F B Mitchell, C A Wilson and C M Price
DCTN 8	Sept 1984	A survey of possible repercussions of the Soviet river reversal programme on Arctic sea-ice. H Cattle
DCTN 9	Oct 1984	A study of a strong stratospheric warming during January 1982 T D A Fairlie
DCTN 10	Nov 1984	An intercomparison of the Met O 11 N45 model and the Met O 20 11-level model on the Cyber 205 J F Dyson
DCTN 11	Nov 1984	Simulation of observed time series of temperature and precipitation over eastern England. D N Reed
DCTN 12	Nov 1984	Forecasts and analyses of the onset of the Southwest monsoon. R Kershaw
DCTN 13	Nov 1984	On modelling the effects of CO_2 on climate. J F B Mitchell
DCTN 14	Dec 1984	Development of a revised longwave radiation scheme for an atmospheric general circulation model. A Slingo and R C Wilderspin
DCTN 15	Jan 1985	Rossby wavetrains in the stratosphere forced by localised disturbances in the troposphere. C J Marks (Oxford), A O'Neill and V D Pope

DYNAMICAL CLIMATOLOGY TECHNICAL NOTES

DCTN 16	Jan 1985	Observing system experiments using the Meteorological Office's 15-level model. R A Bromley
DCTN 17	Feb 1985	Atmospheric general circulation simulation of the climate of the Mediterranean region. H Cattle
DCTN 18	Feb 1985	Diurnal variation and cloud in a general circulation model. C A Wilson and J F B Mitchell
DCTN 19	Feb 1985	The sensitivity of the Saharan region in a general circulation model. W M Cunnington and P R Rowntree
DCTN 20	Mar 1985	The impact of aircraft observations on analyses and forecasts from the operational 15-level model. B R Barwell
DCTN 21	Mar 1985	Non-linear effects on heat-induced circulation of the tropical atmosphere. A E Gill and P J Phlips
DCTN 22	Apr 1985	Simulation of cloudiness with the U.K. Meteorological Office 11-layer AGCM. A Slingo and R C Wilderspin
DCTN 23	Apr 1985	An integrated approach to the representation of cloud processes.
DCTN 24	May 1985	Nuclear Winter: Calculations of the shortwave radiative effects of soot aerosols. A Slingo and P Goldsmith